



## Invited Review

## Current challenges and future developments in magnetic fabric research

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## ARTICLE INFO

## Keywords:

Magnetic fabric  
Anisotropy of susceptibility  
Anisotropy of remanence  
Non-linearity  
Anisotropy correction in paleomagnetism  
Anisotropy modelling  
Field-dependence  
Future development

## ABSTRACT

Magnetic fabrics are used extensively as proxies for mineral fabrics and to correct paleomagnetic data, and have been proposed to quantify pore fabrics. Today's understanding and interpretation of magnetic anisotropy benefits from decades of carefully observing anisotropic mineral and rock properties, establishing and testing empirical relationships between magnetic and mineral fabrics, and developing models to quantify aspects of anisotropy. Each advance was preceded by some puzzling observations, i.e., data that could not be explained based on the models available at the time, e.g., 'oblique' or 'inverse' fabrics, or lithology-dependent anisotropy-strain relationships. These observations led to numerous experimental and numerical techniques designed to characterize magnetic fabrics and determine their origin. Despite the successful application of magnetic fabrics in many structural and tectonic problems, there are still phenomena and measurements that cannot be explained based on today's magnetic fabric theory. With the purpose of fostering future development, I will touch on three main areas where I see challenges in magnetic fabric research: (1) Experimental characterization of magnetic anisotropy, including the non-uniqueness of reported tensors, and non-linearity; (2) anisotropy modelling and defining the carrier minerals and origin of the magnetic fabric; (3) selection of adequate (sets of) anisotropy tensors to correct paleomagnetic data. The description of the challenges presented here will hopefully help define directions for future research, inform those that are new to anisotropy, and advance our field.

## 1. Introduction

Magnetic fabrics are commonly used as a proxy for mineral alignment and therefore a powerful tool in all areas of geology investigating dynamic processes, including the emplacement of igneous rocks, sediment transport and deformation (Borradaile and Henry, 1997; Borradaile and Jackson, 2004; Borradaile and Jackson, 2010; Cañón-Tapia, 2004; Graham, 1954; Hrouda, 1982; Jackson and Tauxe, 1991; Jackson, 1991; Owens, 1974; Owens and Bamford, 1976; Parés, 2015; Tarling and Hrouda, 1993). Less common is the investigation of pore fabrics and prediction of preferred flow directions by measuring anisotropy of magnetic susceptibility (AMS) after a sample has been impregnated with ferrofluid (Parés et al., 2016; Hrouda et al., 2000; Jones et al., 2006; Pfeleiderer and Halls, 1990, 1994). These data, referred to as magnetic pore fabrics, can help predict how fluids migrate through rocks.

A second area where magnetic anisotropy is important, are paleomagnetic, archeomagnetic, and extraterrestrial magnetism studies. Magnetic anisotropy affects the direction and intensity of remanence, as observed in laboratory-deposited/-deformed and natural rocks, or archeomagnetic artefacts (Anson and Kodama, 1987; Fuller, 1960, 1963; Hargraves, 1959; Jackson et al., 1993; King, 1955; Rees, 1961;

Rogers et al., 1979; Tauxe and Kent, 1984). Anisotropy-induced deviations of magnetization directions and intensities have major consequences for paleogeographic reconstructions or the study of the evolution of the geomagnetic field through time (Bilardello and Kodama, 2010a; Gattacceca and Rochette, 2002; Kent and Irving, 2010; Selkin et al., 2000). Therefore, increasingly sophisticated measures of magnetic anisotropy are applied to correct paleo-directions and -intensities (Biedermann et al., 2019a; Bijaksana and Hodych, 1997; Bilardello and Kodama, 2010a; Borradaile et al., 2001; Collombat et al., 1993; Gattacceca and Rochette, 2002; Gattacceca et al., 2003; Hodych and Bijaksana, 1993; Jackson et al., 1991; Kodama, 1997, 2009; Kodama and Sun, 1992; Selkin et al., 2000; Tauxe et al., 2008; Tema, 2009; Werner and Borradaile, 1996).

The use of magnetic fabrics as a tool to rapidly estimate mineral alignment started from careful observations (Fig. 1a): (1) the magnetization acquired by iron spikes depends on the orientation of their long axis during cooling in the geomagnetic field (Gilbert, 1600), and (2) crystal spheres suspended in a magnetic field tend to rotate, and the rotation depends on the orientation of the field with respect to the crystal lattice (Finke, 1909; König, 1887; Plücker, 1858; Stenger, 1888; Tyndall, 1851). These observations illustrate that magnetic properties

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<https://doi.org/10.1016/j.tecto.2020.228632>

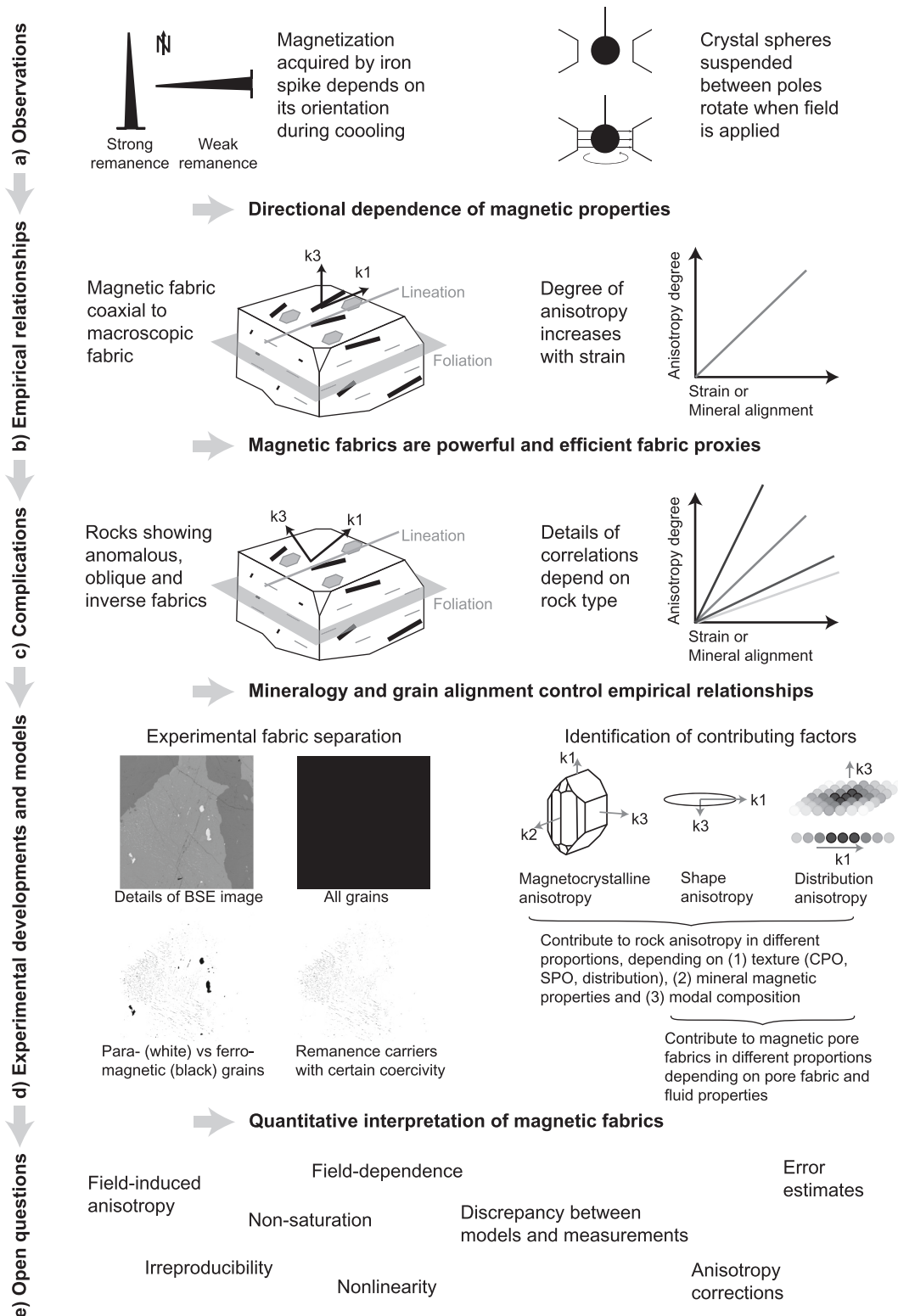
Received 10 August 2020; Received in revised form 14 September 2020; Accepted 16 September 2020

Available online 17 October 2020

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depend on measurement direction, and relate to phenomena known today as shape anisotropy and magnetocrystalline anisotropy, respectively. Comparisons between magnetic fabrics and macroscopic foliation and lineation, or paleoflow and -current directions in rocks revealed

directional relationships (Fig. 1b) (Balsley and Buddington, 1960; Graham, 1954, 1966; Hamilton and Rees, 1970; Ising, 1942; Khan, 1962; Rees, 1965; Stacey, 1960; Taira, 1989), as well as correlations between the degree of magnetic anisotropy and strain (Borradaile, 1988; Cogné



**Fig. 1.** Very brief history of the evolution of our understanding of magnetic fabrics (a-d). New insights and theories were followed by 'complications', i.e., observations that could not be explained by existing theories. Seeking an explanation for these 'abnormal' behaviours eventually led to the development of new theories and an increased understanding about the sources of magnetic anisotropy. (e) The open questions (or complications) discussed in this paper show some of the gaps in our current understanding of magnetic fabrics, and will again lead to new and better theories in the future.

and Perroud, 1988; Hirt et al., 1993; Kligfield et al., 1977; Kligfield et al., 1981). These empirical relationships allow simple geological interpretations of magnetic data, and are still widely used. However, complications were reported in some rocks (Fig. 1c), namely contradicting orientations of magnetic and macroscopic fabrics, (termed inverse, anomalous and oblique fabrics), as well as different AMS-strain relationships for different rock types. These observations highlight the importance of mineralogy and grain size/domain state, and the need to identify carrier minerals for a more sophisticated interpretation (Borradaile, 1987; Henry and Daly, 1983; Housen et al., 1993; Housen and van der Pluijm, 1990; Potter and Stephenson, 1988; Rochette, 1988; Rochette et al., 1999; Rochette et al., 1992).

The desire to understand the sources of these complications and obtain more reliable geologic interpretations from magnetic fabric data, led to major experimental advances and the development of models describing the physical processes that cause magnetic anisotropy (Fig. 1d). Advanced experimental techniques aim at the enhancement or isolation of specific portions of the magnetic fabric or the separation of multiple sub-fabrics (Jackson and Tauxe, 1991; Martín-Hernández and Ferré, 2007). These techniques (1) exploit the different field- and temperature-dependences of para-/dia- and ferromagnetic susceptibilities (Hrouda and Jelinek, 1990; Issachar et al., 2018; Kelso et al., 2002; Martín-Hernández and Hirt, 2004; Pares and van der Pluijm, 2002; Rochette and Fillion, 1988; Schmidt et al., 2007), or (2) are based on specific minerals' abilities to acquire remanence under defined laboratory conditions (Biedermann et al., 2020a; Cox and Doell, 1967; Jackson et al., 1988; Jackson et al., 1989b; Jackson, 1991; McCabe et al., 1985; Potter, 2004; Stephenson et al., 1986). Isolated magnetic sub-fabrics correlate with the crystallographic or shape preferred orientation (CPO and SPO) of constituent minerals, or with the distribution of strongly magnetic grains throughout a rock (Archanjo et al., 1995; Biedermann, 2018, 2019; Cañón-Tapia, 1996, 2001; Chadima et al., 2004; Fuller, 1960, 1963; Grégoire et al., 1998; Grégoire et al., 1995; Hargraves et al., 1991; Hrouda, 1987; Hrouda et al., 1997; Hrouda et al., 1985; Jezek and Hrouda, 2000; Lüneburg et al., 1999; Owens, 1974; Pfeleiderer and Halls, 1993; Siegesmund et al., 1995; Stacey, 1960; Stephenson, 1994). Three factors have been identified to contribute to magnetic fabrics, (1) magnetocrystalline anisotropy in combination with mineral CPO, (2) shape anisotropy in combination with SPO of strongly magnetic grains, and (3) distribution anisotropy arising from magnetostatic interactions between non-randomly distributed strongly magnetic grains. The proportion to which each of them contributes in a specific rock depends on the mineral magnetic properties and the rock's modal composition and texture. Magnetic pore fabrics are controlled by the shape and distribution anisotropies of the ferrofluid in the pore space, which in turn are defined by the pore fabric and ferrofluid properties. Computational models have been developed to help understand and interpret both magnetic fabrics and magnetic pore fabrics (Biedermann, 2019; Cañón-Tapia, 1996; Mainprice et al., 2011; Mainprice and Humbert, 1994; Stephenson, 1994).

Today, quantitative and advanced interpretations of magnetic anisotropy are possible based on our understanding of contributing factors, sophisticated experimental techniques, and numerical models. For example, experimental fabric separation allows the distinction of alteration and deformation events during a rock's history, if the sub-fabrics are carried by different minerals or grain sizes (Almqvist et al., 2012; Aubourg and Robion, 2002; Bilardello and Jackson, 2014; Cioppa and Kodama, 2003; Hirt et al., 1995; Jackson et al., 1988; Kodama and Mowery, 1994; Mattsson et al., 2011; Nakamura and Borradaile, 2001; Raposo and Berquo, 2008; Raposo et al., 2004; Salazar et al., 2016; Sun et al., 1995; Trindade et al., 2001; Trindade et al., 1999). Additionally, numerical models make it possible to study the interplay of fabric contributions that cannot be isolated experimentally, including fabrics controlled by combinations of several paramagnetic minerals. Such models largely facilitate the interpretation of magnetic anisotropy in that they explain the mechanisms leading to anomalous or oblique

fabrics (Biedermann et al., 2018; Kusbach et al., 2019). The experimental advancements in the characterization of magnetic anisotropy also benefit paleomagnetic studies. Remanence anisotropies are preferred to correct for anisotropy-induced deviations in magnetization directions and intensities, because AMS measures the combined response of all grains in a rock, including those that do not carry any remanence (Borradaile et al., 2001; Selkin et al., 2000; Tema, 2009). Better yet are corrections based on the anisotropy of the specific mineral and grain size that carries the characteristic remanence (Borradaile and Almqvist, 2008; Kodama, 1997, 2012; Kodama and Dekkers, 2004). Based on these findings, Biedermann et al. (2019a) proposed improved anisotropy corrections using a set of partial remanence anisotropies, allowing the correction of natural remanence carried by multiple grain populations.

Magnetic fabrics have become a valuable quantitative and robust tool to study numerous dynamic processes. The 'magnetic fabric tool' will be applied to various new geological areas and tectonic settings in the future. These will not be discussed in this study. Rather, I want to focus on our understanding of the tool itself, challenges and limitations of the current experimental and numerical treatment of magnetic anisotropy, and point out open questions and directions for possible future developments (Fig. 1e). The aim of this study is to show some puzzling data that may illustrate gaps in our current treatment and understanding of magnetic anisotropy. The study addresses three main topics within the field of magnetic fabric research: (1) Experimental difficulties, including dependence on experimental settings and non-linearity; (2) mathematical description and models of magnetic anisotropy; and (3) correction of paleomagnetic directions and intensities for the anisotropic acquisition of remanence. Additional challenges will surface (and hopefully be addressed) as our field evolves.

## 2. Laboratory measurements of anisotropy

### 2.1. General description of magnetic anisotropy

Magnetic anisotropy is commonly characterized by measuring magnetic susceptibility or remanence in different directions, or by measuring susceptibility differences in mutually perpendicular planes. An ellipsoid is then fitted to the directional data, which describes the second-order tensor relating the applied field to magnetization (Cogné, 1987; Girdler, 1961; Jelinek, 1977, 1996; McCabe et al., 1985; Stephenson et al., 1986). The tensor's eigenvalues ( $k_1 \geq k_2 \geq k_3$ ) and eigenvectors are referred to as principal susceptibilities and principal susceptibility directions, respectively. Full tensors are derived from directional measurements of susceptibility, in which case  $\frac{k_1+k_2+k_3}{3} = k_{mean}$  with  $k_{mean}$  being the mean susceptibility, and the eigenvalues are often reported in normalized form, with their sum either 3 or 1. Measurements of susceptibility differences lead to deviatoric tensors, in which case  $k_1 + k_2 + k_3 = 0$ . Both sets of eigenvalues are commonly called  $k_i$ , but it can easily be determined which type of tensor they were derived from based on their sum. The full and deviatoric tensors are related by  $\mathbf{k}_{full} = \mathbf{k}_{deviatoric} + k_{mean}\mathbf{I}$ , where  $\mathbf{I}$  is the  $3 \times 3$  identity matrix; bold italic type is used for matrices, and italics for scalar values. Degree and shape of anisotropy are calculated from the eigenvalues, and numerous definitions exist (Hrouda, 1982; Jelinek, 1981, 1984). The anisotropy degree  $P = k_1/k_3$  and shape parameter  $T = (2 * \ln(k_2) - \ln(k_1) - \ln(k_3))/(\ln(k_1) - \ln(k_3))$  are often used in structural interpretations.  $P$  relates to texture strength as long as the mineralogy and mean susceptibility are constant, but needs to be interpreted in combination with  $k_{mean}$  when evaluating the contributions of different minerals to the overall anisotropy in a rock. Both  $P$  and  $T$  are defined solely for full tensors. Here, the mean deviatoric susceptibility  $k' =$

$$\sqrt{((k_1 - k_{mean})^2 + (k_2 - k_{mean})^2 + (k_3 - k_{mean})^2)} / 3 \quad \text{and} \quad \text{shape}$$

parameter  $U = (2 * k_2 - k_1 - k_3)/(k_1 - k_3)$  are used, as they are universally applicable to both full and deviatoric tensors, and because  $k'$  values of different sub-fabrics relate directly to the respective minerals' contributions to a rock's anisotropy.

One important prerequisite for the tensor description of anisotropy is the linear relation between magnetization and field,  $\vec{M} = k\vec{H}$  (Coe, 1966; Jelinek, 1977). Linearity is normally assumed for low-field measurements, e.g. AMS and anisotropy of anhysteretic remanence (AARM) or thermal remanence (ATRM). Conversely, the tensor representation is not strictly correct for high-field applications such as the anisotropy of isothermal remanence (AIRM), because the isothermal remanence increases nonlinearly with field (Coe, 1966). For high-field measurements it is assumed that the magnetization of ferromagnetic grains is saturated and field-independent, whereas the contribution of paramagnetic and diamagnetic minerals to the magnetization increases linearly with applied field (Hrouda and Jelinek, 1990; Kelso et al., 2002; Martín-Hernández and Hirt, 2001; Rochette and Fillion, 1988). Recent observations on pyroxene-hosted magnetite exsolutions with strong shape anisotropy suggest that the  $\vec{M}$  vs  $\vec{H}$  relationship is more complicated in some minerals and rocks (Biedermann et al., 2020b).

The second order tensor representation is convenient, as it allows for direct comparison between magnetic anisotropy and strain markers or other second-order tensor properties such as permeability, diffusivity, and the optical indicatrix. Symmetry constraints on second-order tensor properties permit an initial quality control of single crystal measurements (Neumann, 1985; Nye, 1957). Nonlinearity between magnetization and field complicates the description of magnetic anisotropy. Several options to present such data have been proposed, but our community still needs to reach common agreement on the treatment and representation of these data.

## 2.2. Low-field AMS

Low-field AMS is by far the most commonly employed measure of magnetic fabric. The fields applied are on the order of a few 100 A/m (few 100  $\mu$ T). Paramagnetic and diamagnetic minerals as well as magnetite show linear relationships between  $\vec{M}$  and  $\vec{H}$  in these low fields, i.e., their susceptibility  $k$  is independent of field. Conversely, some minerals including pyrrhotite, titanomagnetite and some but not all hematite samples show field-dependent susceptibilities even in low fields, which is also reflected in their AMS (de Wall, 2000; de Wall and Worm, 1993; Guerrero-Suarez and Martín-Hernández, 2012; Hrouda, 2002, 2011a; Hrouda et al., 2018; Jackson et al., 1998; Volk et al., 2018; Worm, 1991). Processing these measurements with linear AMS theory can result in seemingly negative minimum susceptibilities (Hrouda, 2002). In line with Coe (1966)'s statement that 'a less restrictive analysis than that offered by tensors will be needed', Hrouda et al. (2018) suggested alternative ways of presenting low-field AMS data, e.g. by contour plots of directional susceptibility values. They also concluded, however, that the deviation from second-order tensor symmetry was small enough in their measurements to be neglected. On the other hand, field-dependent and field-independent contributions to low-field AMS can be separated, providing further insight into different anisotropy components (Hrouda, 2009).

In addition to the field-dependence occurring in some minerals, low-field susceptibility and AMS can display frequency-dependence (Hrouda, 2011b), and out-of-phase components (Hrouda et al., 2017). Frequency-dependence serves as grain-size indicator (Dearing et al., 1996; Eyre, 1997), and the out-of-phase susceptibility can isolate the fabric of minerals such as pyrrhotite, hematite, titanomagnetite and superparamagnetic magnetite (Hrouda et al., 2017). Frequency-dependence is expected in superparamagnetic particles and particularly important for magnetic pore fabric studies or for fine-grained rocks and soils. Both field- and frequency-dependence complicate the

comparison of results measured at different experimental conditions.

An aspect of low-field AMS that may largely affect structural interpretations, and thus deserves further attention, is the occurrence of field-induced anisotropy. Field-induced anisotropy results from alignment of domain walls during the treatment with laboratory fields (Bhathal and Stacey, 1969; Halgedahl and Fuller, 1981; Kapicka, 1981; Potter and Stephenson, 1990; Stacey, 1961, 1963; Violat and Daly, 1971), and is mainly observed in samples with high coercivities in some sample sets (Kapicka, 1981), and those with low anisotropy degrees and low median destructive fields in others (Biedermann et al., 2017). An easy way to avoid field-induced effects is to measure low-field AMS prior to any laboratory treatment.

## 2.3. AARM

AARM is viewed as the best room-temperature proxy of the anisotropy that affects natural (partial) thermal remanence (Potter, 2004). Directional remanences are imparted by low DC bias fields applied on top of a decaying alternating field (AF). Applied DC fields span the range from 10  $\mu$ T to 1 mT, and the most commonly used fields are 50  $\mu$ T (self-reported in community survey) and 100  $\mu$ T (published) (Biedermann et al., 2020a). Due to the low bias fields used, anhysteretic remanence is commonly treated as linear with field, and AARM data processed with linear anisotropy theory. However, anhysteretic remanence is not necessarily linear with DC field even in these low field ranges (Sugiura, 1979). Bilardello and Jackson (2014) reported that the AARM fabric orientation, degree and shape of the anisotropy can vary with DC field. Their samples show stronger anisotropy and largest tensor misfits at the lowest DC field used, which they linked to nonlinear remanence acquisition (Fig. 2a). When the relationship between anhysteretic remanence and DC field is nonlinear, the standard treatment of AARM as a susceptibility tensor (the anisotropy of anhysteretic susceptibility, AAS, defined by (Jackson et al., 1989a; McCabe et al., 1985) but more recently termed 'AARM') is not mathematically correct, though it may be a good approximation. To distinguish between linear and nonlinear cases, Bilardello (2020) suggests to use the term AAS and report results in units of susceptibility if linearity is proven, and the term AARM and units of magnetization otherwise.

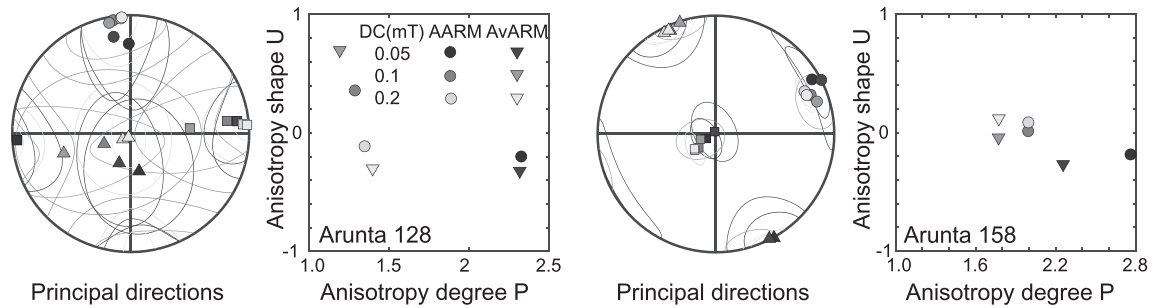
Fig. 2b shows the DC-field dependence of ARM acquired in different coercivity windows for various rock types. The data indicate that the deviation of the ARM-DC field relationship from linearity in a given sample can vary depending on the AF window over which the DC bias field was applied. An additional dependence of this (non)linearity and its variability with AF window on measurement direction cannot be excluded and needs to be investigated further.

Decay-rate dependence of ARMs (Biedermann et al., 2019a; Sagnotti et al., 2003; Yu and Dunlop, 2003) may affect AARMs, and is an additional indication of the complicated physics behind ARM and AARM acquisition. Furthermore, AARM results in coercivity spectra that are directionally dependent. This phenomenon, known as switching-field angular dependence, can lead to overestimation of the AARM in a given coercivity window, because the fraction of grains that is activated by a given field depends on the field direction (Finn and Coe, 2020; Madsen, 2004). Additional phenomena that may affect AARM measurements and need further work or suitable experimental protocols are gyroremanence and rotational remanence resulting from alternating fields (Finn and Coe, 2016; Potter, 2004; Roperch and Taylor, 1986; Stephenson, 1980, 1993).

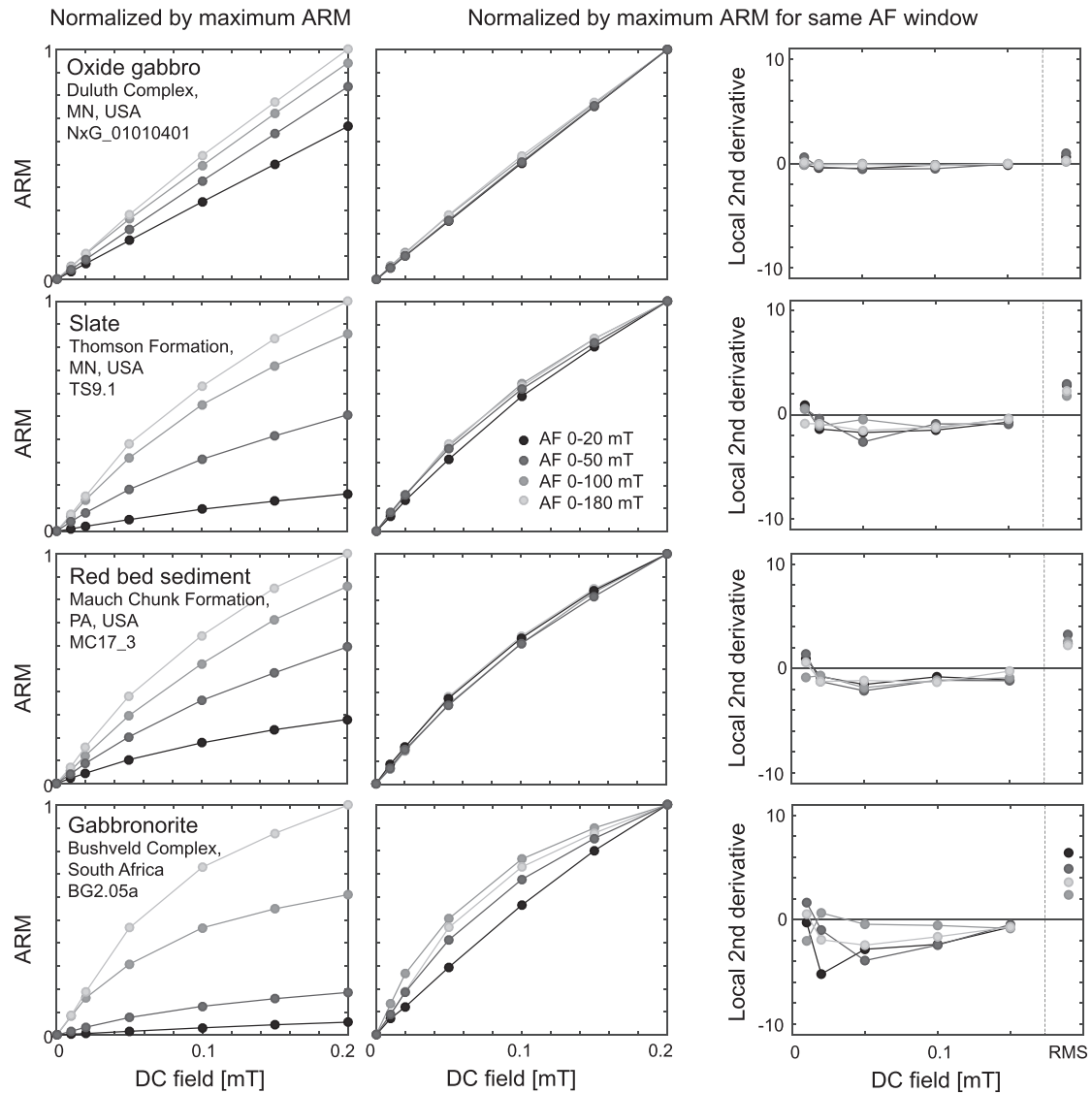
In summary, AARMs have great potential to describe anisotropic acquisition of natural remanence (Potter, 2004), and anisotropy of partial ARMs (ApARMs) can distinguish between grain sizes and shapes (Biedermann et al., 2020a; Jackson et al., 1988; Jackson et al., 1989b). However, despite attempts to model ARMs (Egli, 2006; Egli and Lowrie, 2002), the physics of ARM acquisition is complex and hard to predict. As a consequence, AARM and ApARM tensors are not unique descriptions of magnetic fabrics. More work will be needed to understand DC-field-



a) DC-field dependence of AARM: Granulite, Central Australia, Bilardello &amp; Jackson 2014



b) DC-field dependence of ARM: Various rock types, Biedermann unpublished



**Fig. 2.** (a) DC-field dependence of anhysteretic remanence anisotropy. Different symbols indicate remanences acquired at different DC fields, and using parallel-component (AARM) or full-vector calculations (AvARM). Principal directions shown as lower hemisphere equal area stereoplots where square, triangle and circle correspond to maximum, intermediate and minimum principal anhysteretic susceptibilities, and solid lines in corresponding colors are the 95% confidence ellipses. Redrawn after Bilardello and Jackson (2014). (b) Acquisition of ARM with DC field for various rock types and in different AF windows, showing various degrees of nonlinearity between magnetization and applied field.

and decay-rate dependencies, switching-field angular dependence, gyroremanence and rotational remanence, and to define the tensor measurements that provide the most appropriate correction for paleomagnetic data, or the best description of mineral alignment.

## 2.4. AIRMs

AIRMs are sometimes favoured over AARMs for two reasons: (1) they are less affected by measurement noise in weak samples, and (2) they can target higher coercivity minerals such as hematite, which cannot be reached by AARMs (Bascou et al., 2002; Bilardello and Kodama, 2009; Bogue et al., 1995; Cagnoli and Tarling, 1997; Cox and Doell, 1967; Font et al., 2005; Kodama and Dekkers, 2004; Kovacheva et al., 2009; Potter, 2004; Stephenson et al., 1986; Stokking and Tauxe, 1990; Tamaki and Itoh, 2008). The fields used to impart directional isothermal remanent magnetizations (IRMs) range from 10s of mT to 13 T, implying that the departure from linear field-magnetization relationships is larger than for AMS and AARMs, and the potential effects of analysing AIRMs with tensor mathematics require further study, or alternative descriptions (Coe, 1966).

In addition to non-linearity, the following difficulties may arise in AIRM studies, and need to be investigated in more detail: (1) Directional IRMs are not always reproducible in repeat measurements, especially when pulse magnetizers are used to impart the remanences (Biedermann et al., 2019b; Roberts, 2006). This uncertainty in directional measurements particularly affects the anisotropy calculations in samples with weak anisotropy. (2) Strong directional IRMs cannot be fully demagnetized between measurement directions (Biedermann et al., 2019b), again affecting the anisotropy calculation, and the isolation of partial AIRM fabrics. A multispecimen method has been proposed to circumvent these issues (Bilardello, 2015), but relies on homogeneous samples. (3) Similarly to AARMs, AIRMs may be affected by the angular dependence of the switching-field (Stephenson and Shao, 1994). Tauxe et al. (1990) observed apparent changes in coercivity during AIRM measurements, which they attributed to metastable domains in hematite, another effect that needs further investigation.

## 2.5. ATRMs

Anisotropy of full or partial thermal remanence (ATRM, ApTRM) would be the most adequate description of the anisotropy of a natural TRM (Cogné, 1987). However, chemical alteration is a major limitation for any method involving heating of the sample, and the large number of heating steps required for A(p)TRM experiments make sample alteration even more likely (Potter, 2004). Alternative methods have been developed in the paleointensity community (Shaw, 1974; Walton et al., 1993; Walton et al., 1992; Walton et al., 1996), to minimize sample alteration. These methods may be useful also for anisotropy determination.

Despite the low fields used for ATRM determination (50  $\mu$ T), TRMs are not necessarily linear with field, thus giving rise to the same non-linearity problems as discussed above for AMS, AARM and AIRM (Coe, 1967; Dunlop and Argyle, 1997; Selkin et al., 2007). Finally, cooling rates affect paleointensity determinations (Bowles et al., 2005; Halgedahl et al., 1980; Walton, 1980), and it would be interesting to investigate whether these cooling rate effects have any directional dependence, in which case the anisotropy measured in the laboratory may deviate from the anisotropy controlling natural remanence acquisition.

## 2.6. High-field AMS

High-field AMS is applied to separate paramagnetic and ferrimagnetic anisotropy contributions (Ferré et al., 2004; Hrouda and Jelinek, 1990; Kelso et al., 2002; Martín-Hernández and Hirt, 2001; Richter and van der Pluijm, 1994), and sometimes to separate the hematite/pyrrhotite fabric from the magnetite and paramagnetic fabrics (Martín-

Hernández and Hirt, 2004; Rochette and Fillion, 1988). For torque data, the paramagnetic and ferromagnetic sub-fabrics are separated based on the assumption that the paramagnetic torque increases as a function of field-squared, whereas the ferrimagnetic component is saturated and independent of field. For both contributions, the 20 components, related to second-order tensor anisotropies, are processed (Hrouda and Jelinek, 1990; Martín-Hernández and Ferré, 2007; Martín-Hernández and Hirt, 2001). High-field low-temperature measurements also allow the determination of the diamagnetic contribution to anisotropy (Schmidt et al., 2007). Additional high-field AMS techniques are based on field- and temperature-dependent magnetization measurements on a rotating sample in a cryogenic magnetometer (Rochette and Fillion, 1988), or on hysteresis loops measured in different directions (Ferré et al., 2004; Kelso et al., 2002; Richter and van der Pluijm, 1994). The latter measurements are challenging due to artefacts resulting from sample shape, position, and the spatial sensitivity of the pick-up coils (Kelso et al., 2002), and are thus not discussed further here.

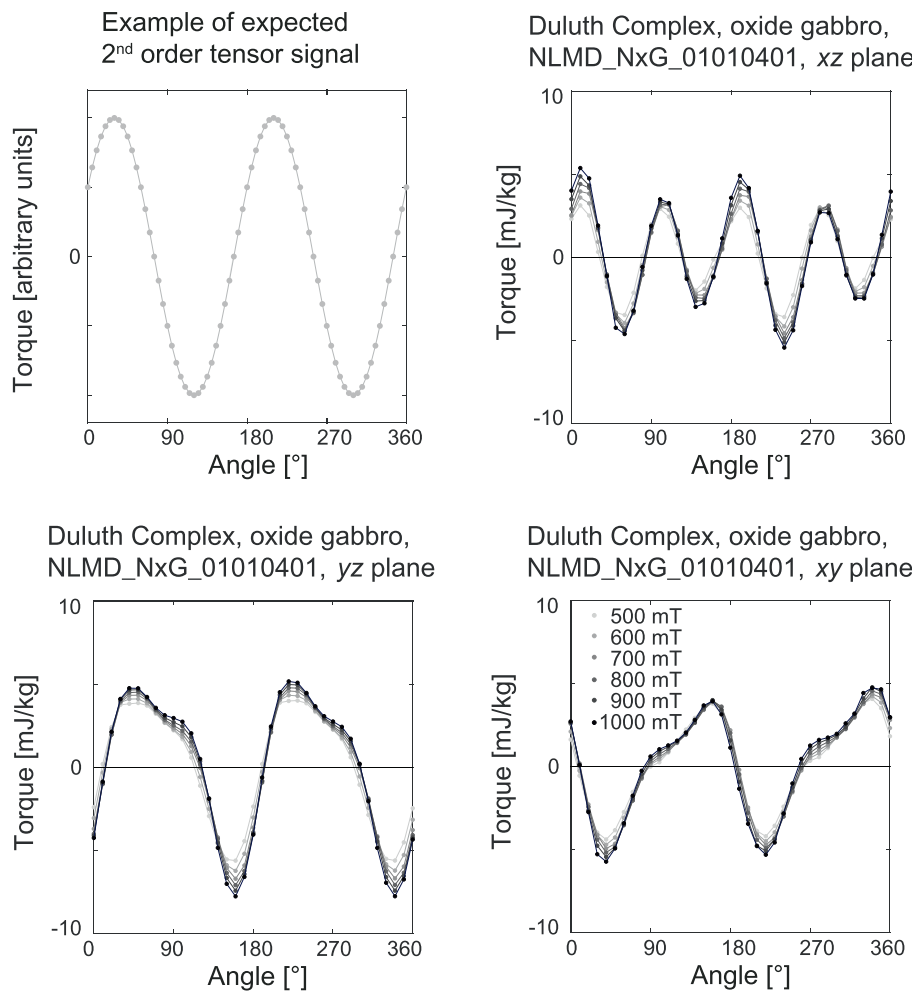
Torque is normally measured in three perpendicular planes, making it possible to assess the validity of the second-order tensor description of the anisotropy. Some torque measurements show higher-order components that are not compatible with second-order tensor anisotropy. This observation has been attributed to magnetocrystalline anisotropy in magnetite, hematite, and pyrrhotite (Flanders and Remeika, 1965; Flanders and Schuele, 1964; Martín-Hernández et al., 2006; Martín-Hernández et al., 2008; Stacey, 1960; Syono, 1965), and recently to strong shape anisotropy in silicate-hosted magnetite exsolutions leading to incomplete saturation of the magnetization (Biedermann et al., 2020b). The reported higher-order components illustrate that second-order tensors are not always sufficient to describe magnetic fabrics, as stated by Coe (1966). Higher-order components and incomplete saturation bear consequences on fabric separation, which usually enforces second-order tensor symmetry. Related effects on structural and tectonic interpretations need to be investigated in future studies. Additional information may be gained from these data if the community finds a way to treat and analyse them adequately. For example, torque signals measured on clinopyroxene crystals may yield information about the aspect ratios and relative proportions of X and Z inclusions of iron oxides they host (Biedermann et al., 2020b) (Fig. 3). One particularly interesting aspect is to determine whether these higher-order components are specific to certain minerals, to a given field range, or other experimental conditions.

# 3. Modelling: Understanding the origin of magnetic anisotropy

## 3.1. Sources of magnetic fabrics

Three main mechanisms have been identified that contribute to magnetic fabrics: magnetocrystalline, shape and distribution anisotropy. These sources, their quantitative mathematical description, and related challenges will be described here in separate sections.

Additional phenomena affecting the magnetic anisotropy of rocks have been reported in experimental studies, and sometimes modelling has been attempted, although they are not commonly included in magnetic fabric interpretations. Field-induced anisotropy is one such phenomenon. A consequence of domain wall alignment in ferromagnetic grains during laboratory field treatment (Bhathal and Stacey, 1969; Halgedahl and Fuller, 1981; Kapicka, 1981; Potter and Stephenson, 1990; Stacey, 1961, 1963; Violat and Daly, 1971), it is not related to any change in CPO, SPO or grain distribution, but to changes of the ferromagnetic grains' internal magnetic structure. Stephenson and Potter (1996) presented a mathematical description of this effect. However, the changes in AMS shape, degree and orientation that are caused by field-induced anisotropy do not provide information on mineral alignment that magnetic fabrics are intended to be a proxy for, and it is favourable to avoid these field-induced artefacts rather than modelling and correcting for them in tectonic applications.



**Fig. 3.** Example of a torque signal resulting from second-order-tensor anisotropy, and torque signals measured in six fields and three mutually perpendicular planes in an oxide gabbro from the Duluth Complex, MN, USA. The oxide gabbro contains pyroxenes with iron oxide exsolutions showing two sets of orientations, controlled by the crystal structure of the pyroxenes. The strong shape anisotropy of these exsolutions results in magnetizations deviating from the field directions, and consequently torque signals that are skewed, asymmetric and contain higher-order components (Biedermann et al., 2020b).

In addition to field-induced domain wall alignment, stress, shock pressure, microstructures and lattice defects have been reported to result in anisotropies unrelated to CPO or SPO (Agarwal et al., 2019; Kapicka, 1988; Nishioka et al., 2007; Park et al., 1988). Possible explanations for these observations are local distortions of the crystal lattice, affecting the crystal field and thus controlling the magnetic moment orientation and magnetic anisotropy. Developments in imaging and direct texture determination, in particular with respect to spatial resolution while maintaining a representative field of view, may make it possible to include these phenomena in future magnetic fabric models. Further sources of AMS or remanence anisotropy that are currently unknown may also be identified.

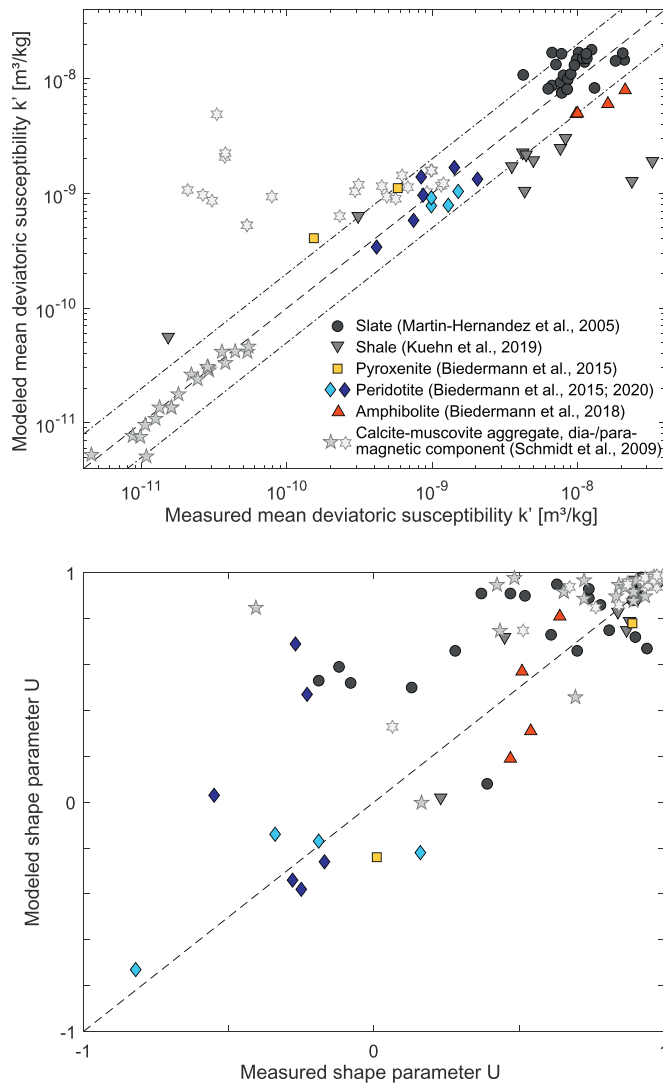
### 3.2. Models based on magnetocrystalline anisotropy and CPO

Anisotropy models based on single crystal anisotropies and CPO are essential to understand the anisotropy contributions of paramagnetic and diamagnetic minerals, and their interplay with each other (Biedermann et al., 2018; Biedermann et al., 2020c; Biedermann et al., 2015; Kuehn et al., 2019; Kusbach et al., 2019; Martín-Hernández et al., 2005; Schmidt et al., 2009). Comparisons between modelled and measured para/diamagnetic anisotropy show mostly good directional agreement, but variability in both anisotropy degree and anisotropy shape (Fig. 4). The good directional agreement is expected if the carriers of anisotropy have been identified correctly, and given the well-characterized single-crystal properties (Biedermann, 2018). Conversely, the mismatch between measured and modelled degree and shape of anisotropy calls for more detailed investigations. Possible explanations for the discrepancies

include (1) sample heterogeneity and the representativeness of the measured mineral orientations, especially when CPO was determined on a single 2D surface; (2) the fact that some minerals in certain orientations are hard to identify with direct texture determining methods (e.g., micas are hard to polish and index in EBSD data, also fine-grained serpentine does not index, but contributes to the magnetic anisotropy), affecting both the modal composition and orientation density function that are used in the models; (3) single crystal susceptibility anisotropy is strongly dependent on Fe concentration (Biedermann, 2018), but models often assume an average single crystal tensor for a given mineralogy due to a lack of chemical analyses. Note that anisotropy degree and shape are more significantly affected by measurement noise than fabric orientation (Biedermann et al., 2013), and analogously, uncertainties in measured CPO or single crystal properties are expected to have a larger effect on modelled anisotropy degree and shape than on fabric orientation. Other possible sources of discrepancy are related to the measurement and fabric separation of magnetic anisotropy discussed above, e.g. when the rocks were measured under different experimental conditions than the single crystal tensors used as model input. Therefore, even though models have contributed largely to our understanding of magnetic fabrics particularly in rocks with composite and complex fabrics, more work is needed until models can unfold their full potential.

### 3.3. Models based on shape anisotropy and SPO

Shape anisotropy arises from self-demagnetization of strongly magnetic non-equidimensional bodies surrounded by less magnetic



**Fig. 4.** (a) Comparison between measured and modelled anisotropy degree, here described by the mass-normalized mean deviatoric susceptibility  $k'$  (Jelinek, 1984). Note log scales on both axes. Dashed line indicates equal anisotropy degree for model and measurement, dash-dotted lines indicate a difference by a factor of two. Different symbols refer to different rock types and studies. (b) Comparison between measured and modelled anisotropy degree, here described by the shape parameter  $U$  (Jelinek, 1981).  $U$  varies from  $-1$  for rotationally prolate ellipsoids to  $+1$  for rotationally oblate ellipsoids. Refer to legend in (a) for an explanation of symbols.

materials, and was first described already in Gilbert (1600) for iron spikes. In rocks, shape anisotropy is mainly relevant for magnetite grains or ferrofluid-impregnated pores. Shape anisotropy can easily be modelled for ellipsoidal bodies (Osborn, 1945; Stoner, 1945), and approximated for other simple geometries such as cylinders and rods (Joseph, 1966, 1967; Sato and Ishii, 1989). For more complicated geometries, the self-demagnetization tensor exhibits spatial variation throughout the body (Joseph, 1976). Therefore, models based on ellipsoidal and cylindrical approximations to grains and pores have been used to investigate the anisotropy of magnetite, or to interpret magnetic pore fabrics (Biedermann, 2019; Biedermann, 2020; Cañón-Tapia, 1996, 2001; Grégoire et al., 1998; Hrouda et al., 2000; Jones et al., 2006). Defining the shape anisotropy of complicated irregular bodies such as magnetite grains or impregnated 3D pore networks would be a logical, though computationally expensive, next step.

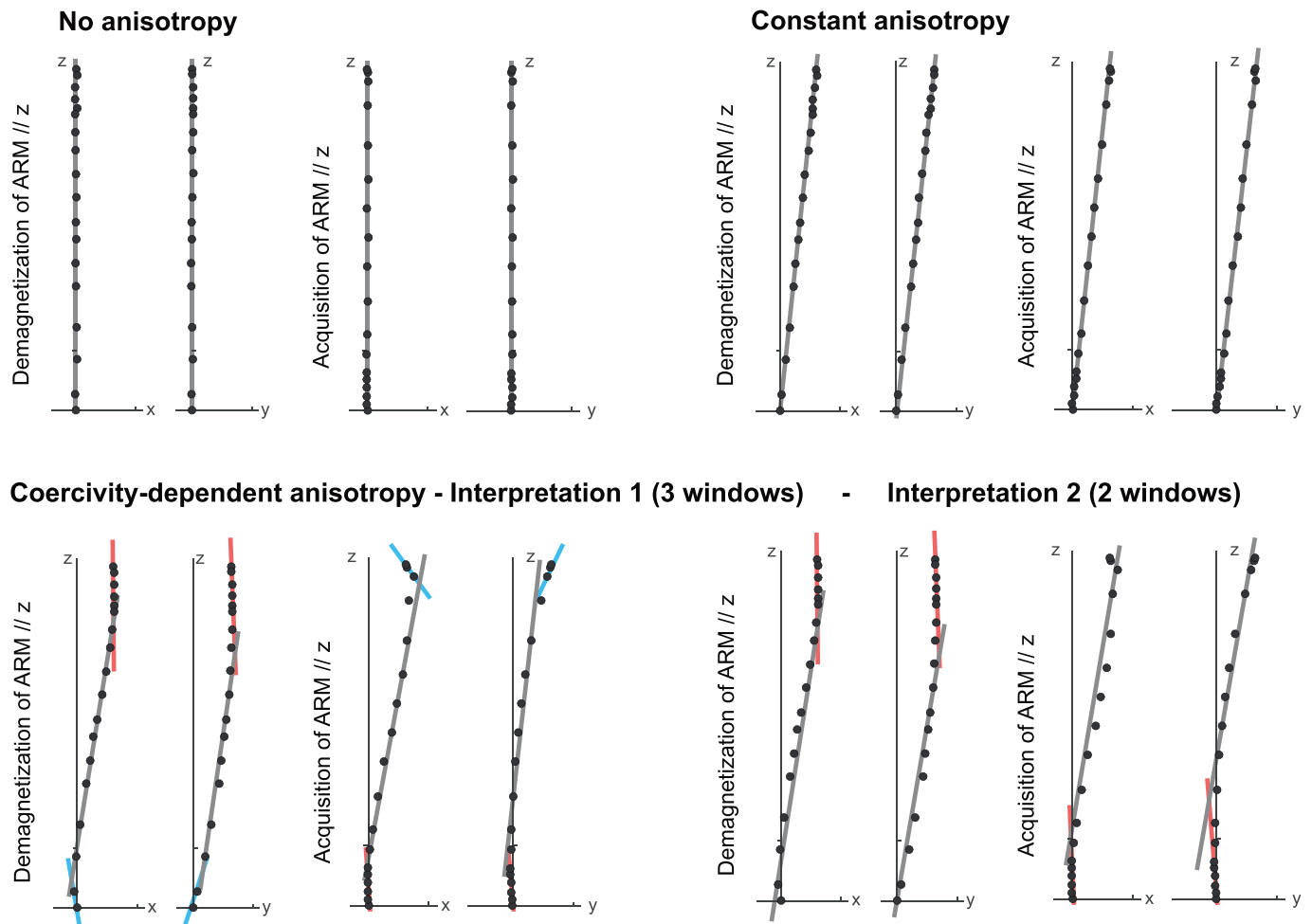
### 3.4. Models based on magnetic interactions and non-uniform grain distribution

Distribution anisotropy is a consequence of the magnetic interactions between strongly magnetic grains or impregnated pores. Its importance with respect to shape anisotropy depends on numerous factors and has long been controversially discussed (Cañón-Tapia, 1996, 2001; Grégoire et al., 1998; Grégoire et al., 1995; Hargraves et al., 1991). Resolution of this controversy was partially hindered by the simplifications necessary when modelling distribution anisotropy, i.e. models existed for equal particles of equal orientation at equal spacing in infinite linear or planar arrangements (Biedermann, 2019; Cañón-Tapia, 1996; Stephenson, 1994). Biedermann (2020) developed a more realistic model, taking into account finite irregular arrangements of particles with different sizes, shapes and orientations. This model presents a step towards solving the controversy, as it allows the prediction of shape and distribution anisotropy for finite irregular particle assemblages more similar to those found in rocks. However, analogous to previous models, it still suffers some simplifications, in particular that particles are approximated by ellipsoidal shapes, and that interactions are calculated based on dipolar secondary fields at particle centres. Therefore, although the model provides results that better match experimental data compared to previous models, adaptations are still necessary before it can be applied to e.g. large irregular 3D networks in pore fabric studies, or to large particles at small spacing, where the interactions and secondary field gradients throughout the particle are strongest.

## 4. Anisotropy corrections in paleomagnetic studies

Reconstructions of plate motions or the magnetic field intensity during the geological past rely on accurate records of field strength and directions. Ideally, a rock's magnetization would be parallel to the field direction and its strength proportional to the field intensity at the time it was magnetized. Deviations from these ideal behaviours are closely related to the magnetic fabric of remanence-carrying grains. The magnetization  $\vec{M} = k\vec{H}$ , where  $k$  describes the anisotropy controlling the characteristic remanence acquisition. Mathematically, the paleofield can then be extracted as  $\vec{H} = \text{inv}(k)\vec{M}$ . Defining the appropriate tensor (or set of tensors)  $k$  is the major challenge. It is commonly agreed that remanence anisotropy targeting the coercivities of the remanence-carrying grains is best suited (Biedermann et al., 2019a; Borradaile and Almqvist, 2008; Kodama, 1997; Kodama and Dekkers, 2004; Selkin et al., 2000). Partial remanence anisotropies can be extracted experimentally for a large range of rock types, and related to grain size or shape (Aubourg and Robion, 2002; Biedermann et al., 2020a; Jackson et al., 1989b; Trindade et al., 2001; Usui et al., 2006), and the AF demagnetization of a laboratory ARM, i.e., a single magnetization event in a known field, can help identify coercivity-dependent remanence anisotropy (Fig. 5; Biedermann et al., 2019a). As shown in that study, it is possible to model the effect of multiple sub-populations of grains with different remanence anisotropies on the magnetization and paleodirectional or paleointensity interpretation. However, defining a set of coercivity windows to measure partial remanence anisotropies that give the most accurate correction is not straightforward. Smaller windows conceivably best represent the gradual changes in magnetization direction shown in Fig. 5. Conversely, small measurement errors have a larger influence on the weaker magnetization in a narrow coercivity window, leading to greater uncertainty in the computed anisotropy tensor. Larger windows yield tensors less affected by noise, but may not sufficiently isolate anisotropies carried by each sub-population, leading to under- or over-corrections. One step towards a more systematic selection of coercivity windows and anisotropy tensors to be determined could be an error propagation that uses the uncertainty in the directional measurements and derived tensor to estimate the potential error range in corrected paleodirections and -intensities, similar to the error





**Fig. 5.** Demagnetization or acquisition of magnetization under known laboratory conditions allows for identification of coercivity-dependent remanence anisotropy. The acquisition of remanence is anisotropic when the magnetization direction deviates from the applied field direction, and if the magnetization direction changes during demagnetization or acquisition, then the remanence anisotropy is coercivity-dependent. The definition of suitable coercivity windows for paleomagnetic anisotropy correction is not straightforward. Data from [Biedermann et al. \(2019a\)](#).

estimation for inclination corrections defined in [Bilardello et al. \(2011\)](#).

An additional difficulty arises when AARMs, AIRMs or ATRMs are used to anisotropy-correct depositional remanent magnetizations (DRMs). Because the processes leading to DRMs (physical rotation of particles) is different from the ARM or TRM acquisition (rotation of magnetic moments in immobile grains), the individual particle anisotropy also has to be defined and corrected for ([Biedermann et al., 2019a](#); [Jackson et al., 1991](#); [Kodama, 2012](#)). Because this is hard to measure and reported values vary (e.g., Table 5 in [Bilardello, 2015](#) for hematite; [Bilardello, 2016](#); [Bilardello and Kodama, 2010b](#); [Kodama, 2009](#)), the correction introduces additional uncertainties, even more so when allowing for multiple sub-populations with different anisotropies. Despite these difficulties, [Tauxe et al. \(2008\)](#) found that AARM-based correction techniques and the elongation/inclination correction method ([Tauxe and Kent, 2004](#)) gave similar results in magnetite- and hematite-bearing sedimentary rocks. Further work will be needed to understand the influence of individual particle anisotropy in anisotropy corrections.

## 5. Conclusions

Magnetic fabrics fit a dual purpose: First, they serve as a proxy for mineral alignment and are thus important tools in tectonic studies and secondly, they allow for correcting paleodirectional and paleointensity data for anisotropic acquisition of remanence. Based on careful

observations and empirical relationships, our community has developed a vast amount of experimental techniques to measure and isolate specific contributions to magnetic fabrics, and an understanding of the main physical sources of anisotropy, and their mathematical description. This knowledge has helped to interpret complex magnetic fabrics, and define sophisticated anisotropy corrections in paleomagnetic applications, with implications for inferred paleogeographic models.

Despite tremendous progress, some aspects of magnetic fabric research still remain puzzling and will need to be investigated further. Major developments in our field have been driven by observations contradicting current theory, and it is with the purpose to foster advancement in magnetic fabric theory that this paper presents aspects needing further work in three areas: (1) Laboratory measurements of magnetic anisotropy, (2) quantitative understanding and modelling of anisotropy, and (3) the determination of adequate anisotropy tensors for paleomagnetic corrections. For the former, the main issues are the dependence of measured magnetic fabrics on experimental parameters, and the nonlinearity between magnetization and field which conflicts with anisotropy theory being developed for linear magnetization-field relationships. Main issues for the second are the discrepancy between modelled and measured anisotropy degree and shape, and the factors affecting anisotropy that cannot be modelled yet, including field-induced anisotropy, shock pressure, microstructure, and lattice defects. Finally, anisotropy corrections will improve when selecting and isolating the anisotropy of the remanence carriers of the relevant

coercivity, but the choice of windows and related error propagation need to be defined. The advances arising from these challenges will make magnetic fabrics even more useful tools in structural, tectonic and fluid migration studies, and I am excited to see the new developments in our field.

## Declaration of Competing Interest

The author declares that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

## Acknowledgements

The article largely benefited from an exchange of ideas with Dario Bilardello who was exploring AARMs and AASs for the IRM Quarterly, volume 30, number 2. I thank Dario for valuable comments and suggestions, and for advertising this article in the Quarterly. Ken Kodama and an anonymous reviewer are gratefully acknowledged for their constructive comments on the manuscript. I am grateful to Zheng-Xiang Li and Tim Horscroft for inviting me to write this review, and to the Swiss National Science Foundation, project 176917, for funding my current research.

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